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S.E. Gryning[1], A.A.M. Holtslag[2], J.S. Irwin[3], B. Sivertsen[4]

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1. Risø National Laboratory, Roskilde, DENMARK
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APPLIED DISPERSION MODELLING BASED ON METEOROLOGICAL SCALING PARAMETERS

S. E. GRYNING*, A. A. M. HOLTSLAG†, J. S. IRWIN‡ and B. SIVERTSEN§

*Risø National Laboratory, Roskilde, Denmark; †Royal Netherlands Meteorological Institute, De Bilt, The Netherlands; ‡U.S. Environmental Protection Agency, Research Triangle Park, North Carolina, U.S.A.; §Norwegian Institute for Air Research, Lillestrøm, Norway.

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Abstract—A method for calculating the dispersion of plumes in the atmospheric boundary layer is presented. The method is easy to use on a routine basis. The inputs to the method are fundamental meteorological parameters, which act as distinct scaling parameters for the turbulence. The atmospheric boundary layer is divided into a number of regimes. For each scaling regime we suggest models for the dispersion in the vertical direction. The models directly give the crosswind-integrated concentrations at the ground, χ_y, for non-buoyant releases from a continuous point source. Generally the vertical concentration profile is proposed to be other than Gaussian. The lateral concentration profile is always assumed to be Gaussian, and models for determining the lateral spread σ_y are proposed. The method is limited to horizontally homogeneous conditions and travel distances less than 10 km. The method is evaluated against independent tracer experiments over land. The overall agreement between measurements and predictions is very good and better than that found with the traditional Gaussian plume model.

1. INTRODUCTION

In practice most of the estimates of dispersion from continuous point sources are based on the Gaussian plume model. A basic assumption for the application of this model is that the plume is dispersed by homogeneous turbulence. However, due to the presence of the surface, turbulence is usually not homogeneous in the vertical direction. It is well established that for a ground level release a Gaussian concentration distribution is not found (Elliott, 1961; Gryning *et al.*, 1983), and especially in very unstable conditions a Gaussian shape is not obtained for elevated releases (Deardorff and Willis, 1975; Briggs, 1985).

The input parameters to the Gaussian plume model are often related to simple turbulence typing schemes or stability classes. A problem with these stability classes is that each covers a broad range of stability conditions, and that they are very site specific. Moreover, these classes are biased toward neutral stability when unstable or convective conditions actually exist (Weil, 1983). The influence of these factors on the calculated ground level concentrations is large (Kretzschmar and Mertens, 1984). To overcome some of the shortcomings of the stability classes, Hanna *et al.* (1977) recommended the use of turbulence measurements to estimate dispersion. Despite these recommendations, most people still use the traditional stability classes (Hanna *et al.*, 1982).

In this paper we briefly review the current concepts of the atmospheric boundary layer for the description of turbulence and dispersion. The idealized boundary layer is divided into a number of regimes, each characterized by distinct scaling parameters (Holtslag and Nieuwstadt, 1986). Based on these parameters, which are easily obtained, we suggest operational dispersion models that contrast with the traditional Gaussian plume model in a number of regimes. The suggested models actually reflect the physical characteristics of the dispersion process within the various regimes of the boundary layer.

The suggested models directly give the crosswind-integrated concentrations at the surface, χ_y, for non-buoyant, non-depositing surface and elevated releases from continuous point sources. The lateral concentration distribution is assumed to be Gaussian. The parameters of the models are briefly summarized and the models are evaluated with independent data from tracer experiments.

2. CHARACTERISTICS OF THE ATMOSPHERIC BOUNDARY LAYER

The atmospheric boundary layer is generally regarded as that part of the atmosphere where the influence of surface friction and heating or cooling is felt. The height of the atmospheric boundary layer during daytime roughly coincides with the height to which pollutants are mixed (the so-called mixing height, h). During night-time stable conditions, h is typically an order of magnitude smaller than the maximum daytime value over land. In this case, h is typically smaller than the height to which the surface radiation inversion extends. Above the sea, the diurnal variation of h is much smaller.

The turbulent structure of the boundary layer can be described with three length scales. These are the height

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above the surface z, the mixing height h, and the Obukhov length L. The height z limits the eddy size to the ground and h limits the vertical extent of the eddies. The Obukhov length L reflects the height at which the contributions to the turbulent kinetic energy from buoyancy forces and from the shear stress are comparable (Obukhov, 1946). The Obukhov length is defined by the surface fluxes of heat $H = \rho c_p \overline{w\theta}_0$ and momentum $u_*^2 = -\overline{uw}_0$ by

$$L = \frac{-u_*^3}{k(g/T_v)\overline{w\theta}_0}, \tag{1}$$

where k is the von Kármán constant, T_v the virtual temperature and g/T_v is the buoyancy parameter. For humid conditions, $\overline{w\theta}_0$ should be replaced by the virtual heat flux (Businger, 1973). The three length scales define two independent non-dimensional parameters. Here we choose z/h, the relative height, and h/L, which can be regarded as a stability parameter for the whole boundary layer.

Figure 1, which is from Holtslag and Nieuwstadt (1986), shows several scaling regimes for the boundary layer. For unstable conditions ($L < 0$) five distinct regimes are identified, in agreement with Panofsky (1978), Nicholls and Readings (1979) and Caughey (1982). The turbulent structures of the surface layer, the free convection layer, and the mixed layer are well understood under horizontally homogeneous and stationary conditions. Less is known for the near-neutral upper layer (Nicholls and Readings, 1979) and the entrainment layer (Deardorff *et al.*, 1980; Driedonks and Tennekes, 1984).

The scaling regimes for the stable boundary layer ($L > 0$) are: the surface layer, the local scaling layer, the z-less scaling layer, the near neutral upper layer and the intermittency layer. The local scaling layer was recently introduced by Nieuwstadt (1984a, b). Presently, no theory is available for the description of the intermittency layer.

The dividing lines between the regions of Fig. 1 are discussed by Holtslag and Nieuwstadt (1986). The turbulent structure of the stable boundary layer has been investigated less than the unstable counterpart, which means that the dividing lines between the stable regimes are more uncertain.

3. VERTICAL DISPERSION

In this section the structure of the turbulence and its implications for the vertical dispersion models are described. Models are proposed, which are based on the specific turbulent structure in the individual regimes of the atmospheric boundary layer. We do not treat dispersion models for the entrainment layer and the intermittency layer. The different dispersion regimes are defined in Fig. 1. The relative height is z_s/h, where z_s is the source height.

(a) *The surface layer* ($-z_s/L < 1 \wedge z_s/h < 0.1$)

The wind shear and the radiational heating or cooling of the surface plays a dominant role for the structure of the turbulence in the surface layer. The effect of stability on the structure is described by Obukhov similarity theory. Nieuwstadt and van Ulden (1978) have shown that the vertical dispersion from a ground-level source in the surface layer is adequately described by K-models. A number of investigations have shown that the diffusivity of matter, K, can be adequately approximated by the diffusivity of heat

$$K = ku_*z/\phi_h(z/L), \tag{2}$$

where ϕ_h is the non-dimensional temperature gradient. The K-model is often solved numerically. Analytical solutions for the crosswind-integrated concentration χ_y, were obtained as early as in the 1930s with power laws inserted for the profiles of wind speed and eddy diffusivity. Part of the pioneering mathematical work is published by Köhler (1933). The analytical solutions

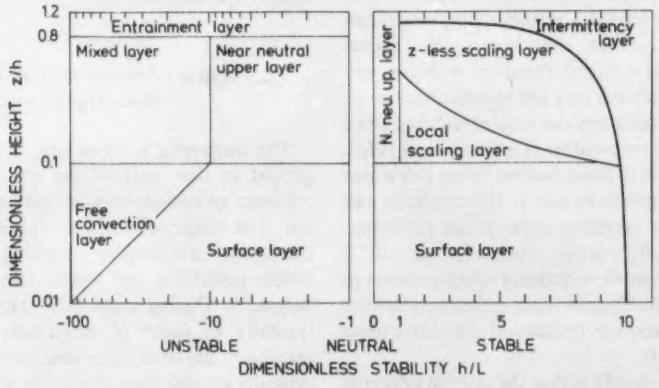

Fig. 1. The scaling regions of the atmospheric boundary layer, shown as function of the dimensionless height z/h, and the stability parameter h/L. A detailed discussion can be found in Holtslag and Nieuwstadt (1986). When used to determine dispersion regions, the dimensionless height is replaced by z_s/h, where z_s is the source height. The boundaries between the regions are given in Section 3.

were elegantly expressed by van Ulden (1978) as:

$$\chi_y(z)/Q = (A/(\bar{z}\bar{u}))\exp\{-(Bz/\bar{z})^s\}, \qquad (3)$$

where Q is the source strength, A and B are functions of the shape parameter s, $\bar{z}$ is the mean height of the plume and $\bar{u}$ is the mean transport velocity of the plume. The definition of $\bar{z}$ and $\bar{u}$ can be found in Gryning *et al.* (1983). Operational methods that can be used to derive these quantities are given in the Appendix for ground level and elevated sources. In (3) deposition is not taken into account.

This model is applicable for $\bar{z}/z_s > 2$. Closer to the source we recommend the use of the models proposed for the near neutral upper layer (Section (b)). When the plume is so far downwind that $\bar{z}$ is substantially bigger than the height of the surface layer, the dispersion is considered to be dominated by the layer above the surface layer, which for unstable conditions is the mixed layer or the near neutral upper layer. We suggest that this is the case when $\bar{z}/h > 0.3$. For elevated sources we recommend use of the models for these layers.

With (3) we obtain a Gaussian concentration profile when $s = 2$. It is characteristic that s is not generally 2 but varies as a function of stability and downwind distance. Under neutral conditions, s approaches 1 far downwind from the source. Under stable conditions, s has an asymptotic limit of 3. The convective limit of s is about 0.5, but the precise value is uncertain. The variation of s has been experimentally verified for near-neutral conditions by Gryning *et al.* (1983) and for stable conditions by Ogawa *et al.* (1985).

(b) *Near neutral upper layer* ($-10 < h/L < 1 \wedge 0.1 < z_s/h < 0.8$–1)

This layer often exists over the sea (Nicholls and Readings, 1979) and is present over land in overcast, windy conditions. The characteristics of dispersion in the near neutral upper layer have not been thoroughly investigated. Because our knowledge is so limited we have retained the Gaussian plume model to calculate the crosswind-integrated concentration at ground-level.

$$\frac{\chi_y}{Q} = \sqrt{\frac{2}{\pi}}\frac{1}{u\sigma_z}\left\{\exp\left[-\frac{1}{2}\left(\frac{z_s}{\sigma_z}\right)^2\right] + \exp\left[-\frac{1}{2}\left(\frac{z_s}{\sigma_z}\right)^2\left(1-\frac{2h}{z_s}\right)^2\right] + \exp\left[-\frac{1}{2}\left(\frac{z_s}{\sigma_z}\right)^2\left(1+\frac{2h}{z_s}\right)^2\right]\right\}. \qquad (4)$$

The estimates of the vertical spread in the Gaussian plume model, σ_z is based on Taylor's (1921) theory for plume dispersion in homogeneous turbulence. The theory relates σ_z to the standard deviation of the vertical wind fluctuations, σ_w. The relation can generally be written

$$\sigma_z = \sigma_w t f_z(t/T_z), \qquad (5)$$

where t is travel time and T_z is the Lagrangian time scale. For practical use in the near neutral upper layer we suggest the following interpolation formula for f_z, which has the correct asymptotic limits for small and large travel times

$$f_z = 1/(1+(t/2T_z)^{1/2}) \qquad (6)$$

in combination with the following values of the time scale T_z

$$T_z = 300\text{ s} \quad \text{for } L < 0,$$
$$T_z = 30\text{ s} \quad \text{for } L > 0.$$

The suggestion for f_z and T_z for $L < 0$ is only marginally different from the formula proposed by Draxler (1976) for unstable conditions. However we prefer this version because T_z for large travel times is identical to the Lagrangian time scale.

Whenever possible, measured values of σ_w should be used in (5). If direct measurements are not available, an estimate of σ_w for $z_s/h < 1$ can be obtained from

$$(\sigma_w/u_*)^2 = 1.5[z_s/(-kL)]^{2/3}\exp(-2z_s/h) + (1.7 - z_s/h) \quad \text{for } L < 0 \qquad (7)$$

and (14) later in this paper for $L > 0$. The equation for σ_w/u_* is based on an empirical model for shear-produced variance by Brost *et al.* (1982) and for buoyancy-produced variance by Bærentsen and Berkowics (1984).

(c) *Mixed layer* ($-h/L > 10 \wedge 0.1 < z_s/h < 0.8$)

The dispersion process in the mixed layer is dominated by the asymmetric structure of the turbulence. Throughout most of the layer, except close to the ground and the base of the inversion, downdrafts occupy a greater area than updrafts. Therefore, a particle released from an elevated non-buoyant source has a higher probability of travelling downward than upward. As a result, close to the source the line of maximum concentration descends until it reaches ground-level. Conversely, a particle released from a non-buoyant ground-level source is found to ascend in such a way that the line of maximum concentration rises. This is due to the strongly inhomogeneous structure of the turbulence near the surface (van Dop *et al.*, 1985). The plume behaviour discussed above has been observed in water tank experiments (Willis and Deardorff, 1978), in wind tunnel simulations (Poreh and Cermak, 1984), in Monte Carlo numerical simulations (de Baas *et al.*, 1986) and in field experiments (Briggs, 1983; 1985). In the mixed layer the relevant quantities are the heat flux $(g/T_v)\overline{w\theta}_0$ and the inversion height h. The characteristic velocity scale is (Deardorff, 1970; Tennekes, 1970)

$$w_* = ((g/T_v)\overline{w\theta}_0 h)^{1/3}. \qquad (8)$$

The dispersion process in the mixed layer can be parameterized with w_* and h. The downwind dimensionless distance is then $X = (w_*/h)(x/u)$, where x/u is the travel time. The dimensionless crosswind-integrated concentration at ground-level is $C_y = \chi_y hu/Q$,

where Q is the source strength. The dimensionless source height is $Z_s = z_s/h$, with z_s the height of the source. Briggs (1985) provided an empirical parameterization among X, Z_s and C_y at ground-level

$$C_y = \frac{0.9X^{9/2}Z_s^{-11/2}}{[Z_s^{-3/4} + 0.4X^{9/2}Z_s^{-9/2}]^{4/3}} + \frac{1}{1 + 3X^{-3/2}Z_s^{1/2} + 50X^{-9/2}}. \quad (9)$$

Figure 2 illustrates the behaviour of (9) for a number of Z_s values. It is characteristic that an X-range exists where the ground level concentration increases with source height. The subunity values of C_y near $X = 2$ are associated with an elevated concentration maximum. It can also be seen that, irrespective of the release height, the plume will be well mixed between the ground and the mixing height beyond $X = 5$. According to Briggs (1985), (9) is valid for non-buoyant plumes provided $0.04 \leqslant Z_s \leqslant 1$, and can thus be applied in a large part of the free convection layer as well. The model validation in chapter 5 reveals that a better fit is obtained when the curve of (9) is shifted towards higher values of X. For practical use we suggest replacing X in (9) by $1.7X$.

(d) *The free convection layer* ($-z_s/L < 1 \wedge z_s/h < 0.1$)

This layer is a matching region of both surface layer and mixed layer scaling. To describe the dispersion from near surface releases, Nieuwstadt (1980) suggested

$$C_y = 0.9X^{-3/2}, \quad (10)$$

which for convenience is expressed in mixed layer scaling parameters. It can easily be shown that (10) implies that C_y is independent of h and u_*, and consequently is applicable for the free convection layer. Comparison with results of the Prairie Grass experiments shows that (10) is valid for $0.03 < X < 0.23$ (Nieuwstadt, 1980).

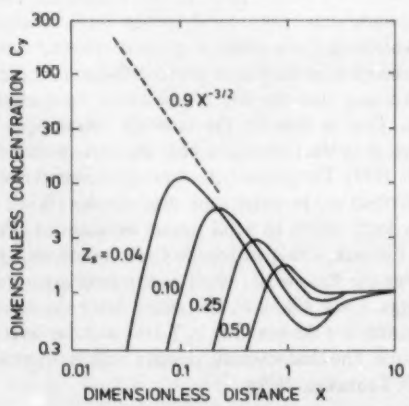

Fig. 2. The solid lines show the dimensionless ground level concentration C_y as function of dimensionless downwind distance X for the mixed layer as given by (9). It is illustrated for a number of dimensionless source heights Z_s. The dashed line shows C_y as function of X for a ground level source in the free convection layer, as described by (10).

No expression similar to (10) exists for elevated sources in the free convection layer. However, Briggs (1985) suggests that (9) adequately describes the dispersion process in the free convection layer provided $Z_s > 0.04$. Figure 2 shows the connection between (9) for mixed layer dispersion and (10) for dispersion in the free convection layer. Note that quantitative agreement between (9), when applied for low sources in the free convection layer and (10) for ground level sources is obtained by shifting the curve of (9) by a factor of ~ 1.7 towards higher values of X.

Holtslag (1984) compared the free convection predictions using (10) with surface layer predictions of (3) (with $s = 1.5$). The comparison was done on the Prairie Grass Data. At $x = 50$ m it was seen that the surface layer model performed better than the free convection layer model, confirming that at $x = 50$ m the plume was still within the surface layer. At $x = 200$ m and $x = 800$ m the predictions of the two models were found to be similar. This suggests that both scaling principles can be used at these distances. The borders between the scaling regions as depicted in Fig. 1, therefore, only give the order of magnitude for the separation of the scaling regions.

(e) *Local and z-less scaling layers* ($h/L > 1 \wedge z_s/h > 0.1 \wedge (h - z_s)/\wedge < 10$)

The turbulence and dispersion in these stable layers have not been so well investigated as in the stable surface layer. Due to the stable stratification the turbulence is suppressed, resulting in a structure completely different from the convective counterpart. Above the stable surface layer, the turbulence scales with the local values of the momentum and heat fluxes (Nieuwstadt, 1984a). Based on the local fluxes, Nieuwstadt formed a local Obukhov length

$$-\wedge = \frac{(-\overline{uw})^{3/2}}{k(g/T_v)\overline{w\theta}}, \quad (11)$$

where $-\overline{uw}$ and $\overline{w\theta}$ are the local fluxes. For horizontally homogeneous and steady state conditions Nieuwstadt (1984a) proposed

$$\wedge/L = (1 - z/h)^{5/4}, \quad (12)$$

where the height of the turbulent layer is given by (Nieuwstadt, 1984b)

$$h = 0.4\left(\frac{u_* L}{f}\right)^{1/2} \quad (13)$$

as originally suggested by Zilitinkevich (1972). Pasquill and Smith (1983) report values in the range 0.2–0.7 for the constant in (13), which strongly stresses that measurements of h are preferred in the calculation. When the turbulent eddies no longer feel the presence of the ground, z is no longer a characteristic scaling parameter for the turbulence and z-less scaling is

appropriate (Wyngaard, 1973). Holtslag and Nieuwstadt (1986) suggest that z-less scaling is applicable for $z/\wedge > 1$ and $(h\text{-}z)/\wedge < 10$. Our knowledge of the characteristics of the vertical dispersion in the local scaling and the z-less scaling layers is very scarce. Therefore we suggest the use of the Gaussian plume model. The vertical spread parameter, σ_z, should be calculated in the same way as for the stable near neutral upper layer, (6), with $T_z = 30$ s. This suggestion is shown by Irwin *et al.* (1985) to be nearly equivalent to the f_z function for stable conditions proposed by Venkatram *et al.* (1984). Direct measurements of σ_w are preferred and strongly recommended. If no measurements are available, however, σ_w can be estimated by

$$(\sigma_w/u_*)^2 = 1.7(1 - z_s/h)^{3/2}, \tag{14}$$

as proposed by Nieuwstadt (1984a) for horizontally homogeneous and stationary conditions.

4. LATERAL DISPERSION

In the previous section the length scales of the turbulence and their implications for the vertical dispersion process were described. Because no restrictions exist in the lateral direction there are no obvious length scales and a similar approach is therefore not possible. The lateral dispersion can be viewed as the combined effect of turbulence of the air and shear in the wind direction. Only the first of these processes will be treated here. The shear effect is not believed to be important within the first 5–10 km from the source, except under very stable conditions (Pasquill and Smith, 1983).

Experiments show that the lateral profile of a plume resembles a Gaussian distribution. (Sivertsen, 1978; Gryning *et al.*, 1978; Nieuwstadt and van Duuren, 1979). Therefore, knowing the crosswind-integrated concentration at the surface, $\chi_y(x, z = 0)$, we can calculate the concentration at the surface at any point using

$$\chi(x, y, 0) = \frac{\chi_y(x, 0)}{(2\pi)^{1/2}\sigma_y} \exp\left\{-\frac{y^2}{2\sigma_y^2}\right\}, \tag{15}$$

where y is the crosswind distance and σ_y is the crosswind spread of the plume. Gryning and Lyck (1984) demonstrate that the best estimate for σ_y is obtained by using a simplified version of Taylor's (1921) formula for plume dispersion. The formula relates σ_y to the standard deviation of the lateral wind fluctuations

$$\sigma_y = \sigma_v t f_y(t/T_y), \tag{16}$$

where f_y is a function of the dimensionless travel time t/T_y and T_y is the Lagrangian time scale for the lateral dispersion.

A comparison of several empirical forms of f_y with field data suggests that the function proposed by Draxler (1976) is best overall (Irwin, 1983; Gryning and Lyck, 1984). Draxler's f_y-function can be written

$$f_y = 1/(1 + (t/2T_y)^{1/2}) \tag{17}$$

and is appropriate for an averaging time of the order of 0.5–1 h. In accordance with Draxler (1976) we recommend that the time scale can be taken as $T_y = 600$ s for elevated sources $(z_s/h > 0.1)$ and $T_y = 200$ s for ground level sources irrespective of the atmospheric stability. However, $T_y = 600$ s should be used also for a ground level source when $\bar{z}/h > 0.1$. Irwin (1983) obtained good results for both elevated and ground level sources with $T_y = 600$ s. Measured values of σ_v at source height are preferred in (16). The measured values should represent an averaging time that corresponds to the averaging time of the concentration values.

However, if measurements of σ_v are not available, it can be estimated for unstable conditions from

$$(\sigma_v/u_*)^2 = 0.35\left(-\frac{h}{kL}\right)^{2/3} + (2 - z_s/h). \tag{18a}$$

This equation is based on an empirical model for shear-produced variance by Brost *et al.* (1982) and for buoyancy-produced variance by Caughey (1982). For stable conditions we propose

$$(\sigma_v/u_*)^2 = 2(1 - z_s/h). \tag{18b}$$

The corresponding averaging time for (18) is roughly ~ 30 min. The expression for σ_v under stable conditions does not include the effect of gravity waves, which for averaging times > 10 min can be very significant, again stressing the importance of measuring σ_v.

5. VALIDATION

The individual models proposed for calculating the atmospheric dispersion in the various regimes of the atmospheric boundary layer have been evaluated in the literature with observations from quite different sources. For low sources the Prairie Grass Data have often been used (Nieuwstadt and van Ulden, 1978; van Ulden, 1978; Horst, 1979; Gryning *et al.* 1983; Holtslag, 1984). For elevated sources a wide variety of data sets exist which cover several source heights, a variety of surface roughness conditions, and a broad range of atmospheric stability (Irwin, 1983; Nieuwstadt and van Duuren, 1978; Briggs, 1985). The method proposed in this paper has also been evaluated using independent data from Copenhagen in Denmark, from Norway, and from the Hanford diffusion grid in Washington.

Gryning and Lyck (1984) describe atmospheric dispersion experiments that were carried out in the northern part of Copenhagen under neutral and unstable conditions. The tracer SF_6 was released without buoyancy from a tower at a height of 115 m, and collected at ground-level positions in up to three crosswind arcs of tracer sampling units. The sampling units were positioned 2–6 km from the point of release.

Tracer releases typically started 1 h before the start of tracer sampling and stopped at the end of the sampling period. The averaging time for tracer sampling was 1 h. The site was mainly residential with a roughness length of ~ 0.6 m. The meteorological measurements taken during the tracer experiments included continuous recording of the three-dimensional wind velocity fluctuations at the height of release, radiosoundings, and time averaged profile measurements of wind and temperature at the 200 m tower where the tracer release took place. From these measurements, values of σ_v, σ_w, the mean wind speed at the release height and the mixing height were determined; the Obukhov lengths were calculated from the profile measurements (Gryning, 1981).

Doran and Horst (1985) describe the SF_6 tracer experiments that were carried out at the Hanford diffusion grid in Washington. The grid is located in a semi-arid region on generally flat terrain. The vegetation consists primarily of desert grasses and 1 to 2 m high sagebrush. The roughness length of the area was 3 cm with a displacement height of 1.4 m. The tracer was released at a height of 2 m and sampled in arcs 100, 200, 800, 1600 and 3200 m from the source. The friction velocity and the Obukhov length were determined by a sonic anemometer and resistance thermometer in four of the experiments, and determined from log–linear fits to observed wind and temperature profiles in the remaining two experiments.

Sivertsen and Bøhler (1985) describe tracer experiments that were carried out in Lillestrøm, a town situated in the southern part of Norway, under stable atmospheric conditions. Only one of the experiments were suited for this type of analysis. The remaining experiments were carried out under meteorological conditions characterized by Richardson numbers bigger than 0.2, which makes any determination of the Obukhov length obscure. The release height of the tracer in this experiment was 10 m; SF_6 was used as tracer. The tracer was sampled near ground-level in arcs 0.7 and 1.5 km from the source with a 15 min averaging time. The site was a residential area with a roughness length of 0.5 m. Meteorological measurements were taken on a 36 m mast. Measurements of wind speed and σ_v were made at the release height (10 m). Also, time averaged profiles of wind and temperature were recorded at the mast. During the experiment a minisonde was launched.

Table 1 shows a summary of the meteorological conditions during the experiments by Gryning and Lyck (1984), Doran and Horst (1985), and by Sivertsen and Bøhler (1985). The relevant scaling regimes for each experiment are indicated. For the experiments in unstable conditions, all of the scaling regimes are covered. For stable conditions, most of the experiments occurred in the surface layer, and only one experiment took place in the local scaling layer.

In Table 2 predicted and the corresponding measured crosswind-integrated concentration for the various experiments are shown. The comparison is illustrated for atmospheric unstable conditions in Fig. 3, and for stable conditions in Fig. 4.

The two experiments in the mixed layer scaling regime qualitatively support the general theory of the behaviour of plumes in the mixed layer (Holtslag *et al.*, 1985; Gryning, 1981) but the measured concentrations are larger than predicted. As discussed by Briggs (1985) and Sivertsen *et al.* (1985) a better fit is obtained when

Table 1. Summary of the meteorological conditions during the experiments that were used for the evaluation of the proposed method for dispersion calculations

Experiment No.	Date	Ref.	Meteorological conditions z_s (m)	h (m)	L (m)	σ_v (ms^{-1})	σ_w (ms^{-1})	u_* (ms^{-1})	u_s (ms^{-1})	z_s/h	h/L	Stability regime
1	20 Sept. 78	GL	115	1980	−46	0.98	0.83	0.37	3.4	0.058	−43	FCL
2	26 Sept. 78	GL	115	1920	−384	1.39	1.07	0.74	10.6	0.060	−5	SL
3	19 Oct. 78	GL	115	1120	−108	0.85	0.68	0.39	5.0	0.103	−10	ML
4	3 Nov. 78	GL	115	390	−173	0.47	0.47	0.39	4.6	0.295	−2.3	NNUL
5	9 Nov. 78	GL	115	820	−577	0.77	0.71	0.46	6.7	0.140	−1.4	NNUL
6	30 Apr. 78	GL	115	1300	−569	2.26	1.33	1.07	13.2	0.088	−2.3	SL
7	27 Jun. 78	GL	115	1850	−136	1.61	0.87	0.65	7.6	0.062	−14	SL
8	6 July 78	GL	115	810	−72	1.35	0.72	0.70	9.4	0.142	−11	ML
9	19 July 78	GL	115	2090	−382	1.71	0.98	0.77	10.5	0.055	−5.5	SL
10	5 Jan. 84	SB	10	90*	60	0.17	—	0.09	0.9	0.11	1.5	LS
11	18 May 83	DH	2	325*	166	—	—	0.40	—	0.0019	2.0	SL
12	26 May 83	DH	2	135*	44	—	—	0.26	—	0.0044	3.1	SL
13	5 June 83	DH	2	182*	77	—	—	0.27	—	0.0033	2.4	SL
14	12 June 83	DH	2	104*	34	—	—	0.20	—	0.0058	3.1	SL
15	24 June 83	DH	2	157*	59	—	—	0.26	—	0.0038	2.7	SL
16	27 June 83	DH	2	185*	71	—	—	0.30	—	0.0033	2.6	SL

SB: Sivertsen and Bøhler (1985); GL: Gryning and Lyck (1984); DH: Doran and Horst (1985); LS: local scaling layer; ZL: z-less scaling layer; SL: surface layer; FCL: free convection layer; ML: mixed layer; NNUL; near neutral upper layer.
* This indicates that h is estimated from (13).

Table 2. Observed and estimated crosswind-integrated concentrations χ_y/Q, and lateral spread parameters, σ_y, for the dispersion experiments of Table 1. "—" indicates that σ_y is not contained in the data set

Scaling regime	Experiment No.	Distance from source (km)	χ_y/Q Observed (10^{-4} sm^{-2})	χ_y/Q Estimated (10^{-4} sm^{-2})	σ_y Observed (m)	σ_y Estimated (m)
Local and z-less scaling layers	10	0.70	440	420	—	
		1.40	150	300	—	
Surface layer	2	2.10	5.38	5.47	239	196
		4.20	2.95	2.85	438	350
	6	2.00	3.96	4.12	—	
		4.20	2.22	2.21	—	
		5.90	1.83	1.78	—	
	7	2.00	6.70	5.84	290	289
		4.10	3.25	2.47	595	520
		5.30	2.23	1.69	786	637
	9	2.10	4.58	5.24	236	243
		4.20	3.11	2.91	460	434
		6.00	2.59	2.47	623	578
	11	0.20	117	261	—	
		0.80	37	78	—	
		1.60	21	43	—	
		3.20	13	25	—	
	12	0.20	367	446	—	
		0.80	129	155	—	
		1.60	91	92	—	
		3.20	72	56	—	
	13	0.20	181	412	—	
		0.80	59	132	—	
		1.60	33	76	—	
		3.20	18	45	—	
	14	0.20	486	628	—	
		0.80	201	215	—	
		1.60	131	128	—	
		3.20	91	78	—	
	15	0.20	424	441	—	
		0.80	105	144	—	
		1.60	86	85	—	
		3.20	66	51	—	
	16	0.20	611	375	—	
		0.80	134	120	—	
		1.60	61	70	—	
		3.20	31	42	—	
Free convection layer	1	1.90	6.48	6.79	254	326
		3.70	2.31	2.55	444	546
Mixed layer	3	1.90	8.20	10.46	184	207
		3.70	6.22	5.68	283	352
		5.40	4.30	3.29	404	471
	8	1.90	4.16	6.03	190	193
		3.60	2.02	3.24	402	330
		5.30	1.52	1.86	580	452
Near neutral upper layer	4	4.00	11.66	7.73	301	221
	5	2.10	6.72	6.20	185	160
		4.20	5.84	4.72	279	280
		6.10	4.97	3.80	376	375

the curve of (9) is shifted towards higher values of X. In the comparison in this paper X was shifted by a factor of 1.7 before using (9).

For atmospheric unstable conditions the mean fractional error between measured and predicted crosswind-integrated concentrations is near -2%. The standard deviation of the fractional errors which describe the ability of the method to predict the variation in the measured χ_y-values is 21 %. In some of the experiments from Copenhagen, the measured crosswind-integrated concentrations were higher than could be predicted by a Gaussian model. This inconsistency is not present in the model proposed here.

In order to compare the skill of the Gaussian model

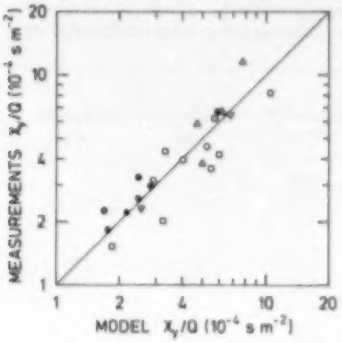

Fig. 3. Estimated vs observed crosswind-integrated concentrations for the experiments that were carried out during unstable conditions. The symbols refer to: ○, surface layer; ●, surface layer when $z/h > 0.3$; Δ, near neutral upper layer; □, mixed layer; ∇, free convection layer.

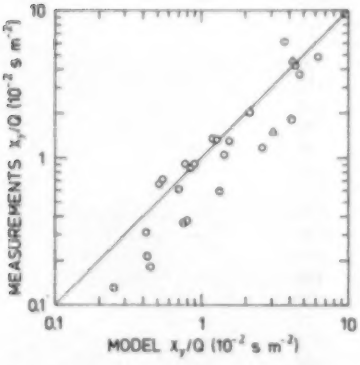

Fig. 4. Estimated vs observed crosswind-integrated concentrations for the experiments that were carried out during stable conditions. The symbols refer to: ○, surface layer; Δ, local scaling layer.

with the method suggested in this paper, all the tracer experiments in the Copenhagen data set were simulated with the Gaussian model that is proposed for the near neutral upper layer. When measured values of σ_w were used as input to the Gaussian model, the mean fractional error between measured and predicted crosswind-integrated concentrations was 6%, and the standard deviation of the fractional error was 34%. When the input to the Gaussian model was based on the recommendations in Turner (1970), which constitutes a kind of industry standard for dispersion calculations, the comparison with the Copenhagen data resulted in a mean fractional error of −4% and a standard deviation of 48%. Thus the method proposed here compares better with the Copenhagen data than both of the methods based on the Gaussian model that were tested above. Gryning and Lyck (1984) report a comparison between measured and predicted crosswind-integrated concentrations from the Copenhagen experiments. Several methods to derive the spread parameters in the Gaussian model were tested. These include methods based on a stability classification of the atmospheric stability and methods based on measured wind variances. The method proposed in this paper is better than any of the methods that were tested by Gryning and Lyck (1984) with regard to the mean as well as the standard deviation of the fractional error.

For atmospheric stable conditions, the methods proposed in this paper work well on most of the experiments. However, in experiments 11 and 13, the predicted crosswind-integrated concentrations are a factor of two greater than the measured ones, which illustrates the difficulties that are encountered when dealing with stable flows. In the comparison we omitted the data from the 100 m arc in the Hanford diffusion grid because the sagebrush close to the source is expected to behave as roughness elements, and the effect therefore is described poorly with a displacement length.

Table 2 gives observed and predicted values of the lateral spread σ_y, based on data from the Copenhagen experiments. The mean fractional error is −4% and the standard deviation of the fractional error is 16%. Figure 5 illustrates the comparison. The ability of the method to predict σ_y under stable conditions was not tested because information on σ_y is not contained in this data set.

6. DISCUSSION

In the paper we have reviewed the scaling arguments for the dispersion of non-buoyant, non-depositing plumes in the turbulent boundary layer. Based on these arguments we related the dispersion of plumes to the turbulent state of the boundary layer. We described the dispersion process in the various scaling regimes of the

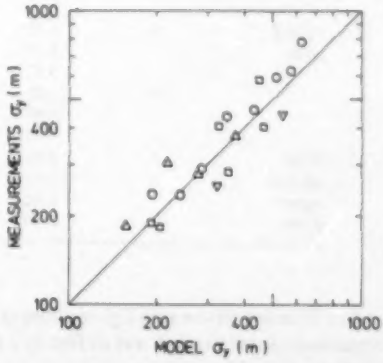

Fig. 5. Estimated vs observed values of σ_y for the experiments that were carried out during unstable conditions. The symbols refer to; ○, surface layer; Δ, near neutral upper layer; □, mixed layer; ∇, free convection layer.

boundary layer in terms of the dimensionless quantities h/L and z_s/h. The method is limited to horizontally homogeneous conditions and travel distances less than 10 km.

In general the vertical dispersion was found to be other than Gaussian. Therefore, we recommend the use of techniques that directly characterize the cross-wind-integrated concentration at the surface, χ_y, whenever possible. For the lateral dispersion a Gaussian distribution is assumed, and the input parameters are suggested. Knowing χ_y, the ground level concentrations can be calculated.

We have compared the suggested models for χ_y with independent data from Copenhagen (Gryning and Lyck, 1984), the Hanford diffusion grid (Doran and Horst, 1985), and Norway (Sivertsen and Bøhler, 1985). It was shown that the overall agreement between estimates and measurements was very good and, under unstable atmospheric conditions, better than that obtained with the Gaussian plume model. Under unstable conditions the mean fractional error between measured and predicted cross-wind-integrated concentrations was -2%. The standard deviation of the fractional errors, which describes the ability of the method to predict the variation in the measured χ_y-values, was 21%.

For convective conditions a quantitatively better agreement with the data is obtained when the curve of (9) is shifted towards higher values of the non-dimensionalized distance X. As can be seen from Fig. 2 such a procedure would give a better agreement with the free convection prediction of (10) for surface sources as well.

The evaluation of the models of this paper was performed with data from tracer experiments where measurements of the meteorological input parameters were available. Irwin *et al.* (1985) discuss instrument requirements for this purpose. The derivation of the meteorological parameters that are needed for air pollution models from routine measurements, is reviewed by van Ulden and Holtslag (1985) and Irwin *et al.* (1985). Wilczak and Phillips (1984) show that the meteorological parameters can be estimated with an accuracy of 10–30% during daytime.

We have not suggested dispersion models for the entrainment and the intermittency layers because the dispersion in these layers is poorly understood and has not been well investigated. Neither have we suggested models to treat the effect of plume buoyancy, whose effect is known to be pronounced. However, the field of plume rise is presently undergoing rapid development (e.g. Briggs, 1985) and is under re-evaluation. Therefore, it is reasonable not to suggest any applied models for the treatment of buoyancy effect because such models will inevitably turn out to be premature. A special complication in the use of the suggested methods arises at the border between the regimes because crossing of the border results in a jump in the calculated concentration.

Improvements in the models proposed here may be introduced by continuous description of the vertical dispersion between the various scaling regions of the boundary layer. Moreover, methods to estimate the effect of buoyancy in the various scaling regimes have to be included when available.

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APPENDIX

In Section 3a we introduced Equation (3) for the calculation of χ_y/Q from the mean height of the plume $\bar{z}$, the shape parameter s, the mean transport velocity of the plume $\bar{u}$ and the functions A and B. Here we will summarize equations for the calculation of these quantities.

The mean height of the plume $\bar{z}$ depends on travel distance x, roughness length z_0 and Obukhov length L. Their relation can be approximated by (van Ulden, 1978.)

$$x + x_0 = (\bar{z}/k^2)[\ln(c\bar{z}/z_0) - \Psi(c\bar{z}/L)] \times [1 - pa_1\bar{z}/(4L)]^{-1/2} \quad \text{for } L < 0, \tag{A1a}$$

$$x + x_0 = (\bar{z}/k^2)[\{\ln(c\bar{z}/z_0) + 2b_2p\bar{z}/(3L)\} \times \{1 + b_1p\bar{z}/(2L)\} + (b_1/4 - b_2/6)p\bar{z}/L] \quad \text{for } L > 0, \tag{A1b}$$

where k is the von Kármán constant and x_0 is an integration constant that accounts for the height of the source (see below). The coefficients p and c in (A1a, b) depend on s. For practical applications we propose $p = 1.55$ and $c = 0.6$. The remaining constants and the Ψ functions are related to the actual choice of ϕ_m and ϕ_h functions (Table A1). Gryning *et al.* (1983) and Holtslag (1984) used the ϕ_m and ϕ_h functions suggested by Dyer (1974) with $a_1 = a_2 = 16$, $b_1 = b_2 = 5$ and $k = 0.41$.

The shape parameter s is approximated by (Gryning *et al.*, 1983):

$$s = \frac{1 - a_1c\bar{z}/(2L)}{1 - a_1c\bar{z}/L} + \frac{(1 - a_2c\bar{z}/L)^{-1/4}}{\ln(c\bar{z}/z_0) - \Psi(c\bar{z}/L)} \quad \text{for } L < 0 \tag{A2a}$$

and

$$s = \frac{1 + 2b_1c\bar{z}/L}{1 + b_1c\bar{z}/L} + \frac{1 + b_2c\bar{z}/L}{\ln(c\bar{z}/z_0) + b_2c\bar{z}/L} \quad \text{for } L > 0. \tag{A2b}$$

It is seen that s is a function of $\bar{z}$, z_0 and L.

The mean transport velocity of the plume $\bar{u}$ is given by (van Ulden, 1978).

$$u = u_*/k\,\{\ln(c\bar{z}/z_0) - \Psi(c\bar{z}/L)\} \quad \text{for } L < 0 \tag{A3a}$$

$$u = u_*/k\{\ln(c\bar{z}/z_0) + b_2\bar{z}/L\} \quad \text{for } L > 0. \tag{A3b}$$

Table A1. General expressions of the ϕ_h and ϕ_m functions that define the coefficients a_1, a_2, b_1 and b_2. The corresponding Ψ-functions are given. The Ψ-function for unstable conditions is an approximation suggested by Jensen *et al.* (1984)

For $L < 0$	For $L > 0$
$\phi_h(z/L) = (1 - a_1z/L)^{-1/2}$	$\phi_h(z/L) = 1 + b_1z/L$
$\phi_m(z/L) = (1 - a_2z/L)^{-1/4}$	$\phi_m(z/L) = 1 + b_2z/L$
$\Psi(z/L) = \phi_m^{-1} - 1$	$\Psi(z/L) = -b_2z/L$

The functions A and B are given by (Pasquill and Smith, 1983):

$$A = s\Gamma(2/s)/[\Gamma(1/s)]^2 \tag{A4}$$

$$B = \Gamma(2/s)/\Gamma(1/s), \tag{A5}$$

where Γ is the gamma function.

If the gamma function is not readily available on the computer, a useful approximation is

$$\Gamma(x) = 1 + \sum_{n=1}^{8} b_n(x-1)^n, \quad 1 \leq x \leq 2, \tag{A6}$$

where

$b_1 = -0.577191652$ $\quad b_2 = 0.988205891$
$b_3 = -0.897056937$ $\quad b_4 = 0.918206857$
$b_5 = -0.756704078$ $\quad b_6 = 0.482199394$
$b_7 = -0.193527818$ $\quad b_8 = 0.035868343.$

From (A1)–(A5) and (3) the concentration profile can be determined. The input parameters for the model are z_0, L and u_*. The roughness length z_0 should be representative of the surrounding area, as discussed by Davenport (1960) and Wieringa (1980).

In practice, the computation with the above equations starts with (A1). First x_0 is computed by putting $\bar{z}$ equal to the source height z_s and $x = 0$. For a ground level source $x_0 = 0$. Then we can determine $\bar{z}$ iteratively from (A1) for any downwind distance x. With $\bar{z}$ we solve for s and $\bar{u}$ using (A2) and (A3). Finally, A and B are calculated with (A4) and (A5), then (3) is evaluated for a given height z in the surface layer.

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